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ABSTRACT

Numerical solutions of mutually coupled time-dependent equations of continuity, momentum and energy balance are presented to illustrate the effect of eddy turbulence on the neutral composition and temperature of the lower and upper thermosphere. The procedure adopted in this paper eliminates the necessity of making ad-hoc assumptions about the conditions at the turbopause level and allows the study of the two regions in a self-consistent manner. From the illustrative examples comprising parametric changes in the eddy diffusion coefficient, the specific roles of eddy turbulence in the development of theoretical models of the thermosphere are discussed.

INTRODUCTION

In developing theoretical models of the upper atmosphere, it is customary to divide the thermospheric region into two domains with a boundary in the altitude range of 100-120 km. This boundary is usually identified with the turbopause level which by definition is the altitude at which the eddy and molecular diffusion coefficients become equal. The two regions, therefore, are distinguished according to the relative importance of molecular and eddy transport. Because of the differences in the geophysical conditions of the two regions, their studies are pursued independently and the link between the two regions is contrived by making ad-hoc assumptions about the boundary values at the turbopause level. The boundary values thus chosen are not necessarily consistent with the equations of continuity, momentum and energy transport of this region and are, in fact, among the most uncertain elements in the development of theoretical models.

An understanding of the nature of variation of neutral composition and temperature in the altitude region of the turbopause is needed not only for the general understanding of the thermospheric behavior but also for the understanding of a large number of ionospheric phenomena which are strongly influenced by changes in the neutral atmosphere. In a series of papers, Chandra and Herman [1969], Chandra and Stubbe [1971] and Chandra et al. [1972] have shown how the changes in neutral composition at the turbopause trigger a complex chain of events, which not only affect the distributions of both the ionized and neutral

constituents and their temperatures, but also the emission rates of OI 6300 Å during a SARARC event. The concept of changing neutral composition has been introduced in several studies for explaining the seasonal and magnetic storm related changes in the ionosphere [King, 1967; Duncan, 1969; Fatkullin, 1973]. Some of these variations in the lower thermosphere have also been inferred from OGO-6 neutral mass spectrometer measurements [Reber et al., 1973]. On a day-to-day basis, there is considerable variability in the conditions of the lower thermosphere. This has been inferred from several rocket measurements of density and temperature [Spencer et al., 1970; Smith et al., 1972]. Although no clear pattern about the nature of these variations has been established yet, it is obvious that the conditions in the lower thermosphere are not invariant as has been assumed in the development of several atmospheric models of the sixties [CIRA 1965, U.S. Standard Atmosphere 1966].

As a first step towards understanding the nature of variations in the altitude region of the turbopause, it is important that we treat the thermosphere as a single entity and solve the relevant equations of the two regions in a self-consistent manner. This will eliminate the necessity of using turbopause as a 'lid' separating the two regions of the thermosphere. We must, in addition, understand the nature of eddy turbulence in the lower thermosphere which is a dominantly controlling factor for both the composition and thermal structure of this region. Unfortunately, no simple way is available at present to theoretically relate thermospheric turbulence in terms of basic geophysical parameters. The

effect of eddy turbulence is, therefore, studied by defining eddy diffusion and eddy conduction fluxes, analogous to molecular diffusion and molecular conduction fluxes of particles and energy, respectively [Colegrove et al., 1965, 1966; Johnson and Wilkins, 1965; Johnson and Gottlieb, 1970]. The eddy diffusion coefficient which determines these quantities is estimated from the sodium vapor trail experiments [Blamont and de Jager, 1961; Zimmerman and Champion, 1963; Justus, 1967; Zimmerman et al., 1971]. Its values have been measured over a wide range and it has not yet been possible to establish a clear pattern for its latitudinal, diurnal, seasonal and height variations. At present, therefore, the effect of eddy diffusion can be best studied by parameterizing the eddy diffusion coefficient, and estimating it indirectly by comparing the observational data with theoretical calculations. This procedure has been adopted by several workers [Ivel'skaya et al., 1970; Shimazaki and Laird, 1970; Shimazaki, 1967, 1971; George et al., 1972]. In most of these studies, however, the equations of continuity are solved with the temperature assumed as a prefixed or known parameter. Since changes in eddy diffusion coefficient also affects the thermal structure significantly, it is clear that the effect of eddy mixing is much more complex than revealed by these calculations [Iwasaka, 1973].

The purpose of this paper is to discuss some of these complexities and their physical implications from the simultaneous solutions of the equations of continuity and energy transport for the thermospheric region. These calculations lead to the development of self-consistent models of the neutral composition and

temperature of the lower and upper thermosphere and allow the study of specific roles of eddy turbulence in a wide variety of geophysical phenomena.

BASIC EQUATIONS

Equations of Continuity

The neutral constituents considered in this study are O, O_2, and N_2, the main constituents contributing to the energetics of the thermosphere in the altitude range of 85-450 km. In deriving their distributions, we have assumed N_2 to be in diffusive equilibrium throughout the above altitude range. The distribution of N_2 is, therefore, uniquely determined from temperature distribution which is obtained by solving the energy balance equation. The only motion attributed to N_2 is the 'breathing' motion caused by the expansion and contraction of the atmosphere. The distribution of N_2 and its velocity are thus given by the following expression:

$$[N_2] = \frac{[N_2]_o T_o}{T} \exp\left[- \int_{z_o}^{z} \frac{dz}{H_{N_2}} \right] \tag{1}$$

$$V_{N_2} = \frac{T}{g} \int_{z_o}^{z} \frac{g(z')}{T^2(z')} \frac{\partial T(z')}{\partial t} dz' + H_{N_2} \frac{\partial}{\partial t} (p_{N_2})_o \tag{2}$$

where the suffix o corresponds to the value at the lower boundary z_o (= 85 km).

Here

$$p_{N_2} = \text{partial pressure of } N_2 = [N_2]\, kT \tag{3}$$

$$H_{N_2} = N_2 \text{ scale height} = \frac{kT}{m_{N_2} g} \tag{4}$$

where g is the acceleration due to gravity, k the Boltzmann constant, m_{N_2} the molecular weight of N_2 and T the temperature.

The assumption of diffusive equilibrium for O and O_2 is valid only in the upper thermosphere ($z \gtrsim 120$ km). In the lower thermosphere their distributions are controlled by chemical, photochemical and transport processes and must be determined by solving the appropriate equations of continuity as follows:

$$\frac{\partial n_i}{\partial t} = q_i - L_i - \frac{\partial}{\partial z}(n_i V_i), \quad (i = 1, 2) \tag{5}$$

where q_i and L_i are production and loss rates, V_i the vertical transport velocity and the suffixes $i = 1$ and 2 designate O and O_2, respectively.

The terms q_i and L_i arise from photodissociation of O_2 [in the Schumann-Runge Continuum ($\lambda = 1250$-1700 Å) and Schumann-Runge Band ($\lambda = 1700$-2100 Å)] and from 3-body recombination and are calculated here in the same way as discussed by Shimazaki [1971]. The velocity V_i in eqn. (5) consists of 3 terms: (i) the velocity V_{N_2} of the major constituent N_2, (ii) molecular diffusion velocity V_{iD}, and (iii) the eddy diffusion velocity V_{iK}, relative to N_2. Thus

$$V_i = V_{N_2} + V_{iD} + V_{iK} \tag{6.a}$$

The vertical mass motion of the neutral gas is thus given by the mean value:

$$V_z = \frac{\sum_{i=1}^{3} \rho_i V_i}{\rho}, \quad \rho_i = m_i n_i, \quad \rho = \sum_{i=1}^{3} \rho_i \tag{6.b}$$

where the suffixes $i = 1, 2$, and 3 denote O, O_2 and N_2, respectively. Here V_{N_2} is given by equation (2) and V_{iD} and V_{iK} are given by the following expressions:

$$V_{iD} = - D_{i3}\left[\frac{1}{n_i}\frac{\partial n_i}{\partial z} + \frac{1}{T}\frac{\partial T}{\partial z} + \frac{1}{H_i}\right] \tag{7.a}$$

$$V_{iK} = - K\left[\frac{1}{n_i}\frac{\partial n_i}{\partial z} + \frac{1}{T}\frac{\partial T}{\partial z} + \frac{1}{\overline{H}}\right] \tag{7.b}$$

where D_{i3} and K are molecular and eddy diffusion coefficients relative to N_2, respectively, H_i is the respective scale height and $\overline{H}$ is the average scale height which approximately corresponds to the scale height of N_2. The coefficient D_{i3} is calculated from the expression of a simple binary diffusion coefficient as follows [Chapman and Cowling, 1952]:

$$D_{i3} = \frac{3}{8\,\sigma_{i3}^{\,2}\left\{n_i + [N_2]\right\}}\left[\frac{kT\left\{m_i + m_{N_2}\right\}}{2\pi\, m_i\, m_{N_2}}\right]^{1/2}, \quad (i = 1,\ 2) \tag{8}$$

where σ_{i3} is the collision diameter ($\sigma_{13} = 2.4 \times 10^{-8}$cm, $\sigma_{23} = 3.4 \times 10^{-8}$cm) [Lettau, 1951].

An expression for K in terms of simple geophysical parameters is not available, as mentioned earlier. Its functional form, therefore, must be described in terms of assumed parameters. In this paper, we have adopted a model of K(z) given by Shimazaki [1971], which basically relates the observational data in terms of a few simple parameters. Denoting the maximum value of eddy coefficient by K_M, and the corresponding height by z_m, we write:

$$K(z) = \begin{cases} K_M \exp\left[-S_1\,(z - z_m)^2\right], & z \geq z_m \\ (K_M - K_o)\exp\left[-S_2\,(z - z_m)^2\right] + K_o \exp\left[S_3\,(z - z_m)\right], & z \leq z_m \end{cases} \tag{9}$$

where S_1, S_2 and S_3 are shape parameters and K_o is a constant. With suitable

choice of these parameters, the eddy diffusion coefficient can be made to fall off rapidly above the height of the maximum, simulating the condition of rapid cessation of turbulence above the turbopause.

Energy Balance Equation

The energy equation for determination of thermal structure above 120 km has been discussed in detail by Chandra and Sinha [1973]. In this paper this equation has been modified to include the terms arising from eddy transport and 3-body recombination. The basic form of energy balance equation solved in this paper is as follows:

$$\sum_{j=1}^{3} \left(Cp_j \, \rho_j \, \frac{dT}{dt} - \frac{dp_j}{dt} \right) = Q_i + Q_d + Q_r + Q_c + Q_e - Q_\ell \qquad (10)$$

with

$$\frac{d}{dt} = \frac{\partial}{\partial t} + V_z \frac{\partial}{\partial z} ,$$

j = 1, 2, 3 for O, O_2, N_2, respectively, and Cp_j is the corresponding specific heat at constant pressure.

The terms on R.H.S. of eqn. (10) represent the contributions from various sources and sinks defined as follows:

Q_i = Thermal energy from photoelectrons produced by photo-ionization of O, O_2 and N_2 in the EUV range of solar spectrum.

Q_d = Energy available from photodissociation of O_2 in the Schumann-Runge continuum and Schumann-Runge band.

Q_ℓ = Heat loss arising due to infrared radiation (63μ) emitted by thermally excited atomic oxygen.

The expressions for Q_i, Q_d and Q_ℓ are given in a previous paper [Chandra and Sinha, 1973] and are calculated here in exactly the same way. Q_r is the energy input rate due to 3-body recombination of O:

$$O + O + M \rightarrow O_2^* + M \tag{11.a}$$

$$O_2^* \rightarrow O_2 + h\nu_H \tag{11.b}$$

The recombination of O normally results in the production of O_2 in the excited state (denoted by the superscript *) and its subsequent de-excitation through emission in the Herzberg band (frequency ν_H). The recombination energy is calculated from the following expression:

$$Q_r = \alpha N [O]^2 (\epsilon_d - \epsilon_H) \tag{12.a}$$

where ϵ_d is the dissociation energy of O_2 (=5.1 ev), ϵ_H the average energy of the Herzberg band (=4.25 ev), N the total number density and α the rate coefficient. α has a strong temperature dependence and is given by [Campbell and Thrush, 1967].

$$\alpha = 3 \times 10^{-33} \left(\frac{T}{300}\right)^{-2.9} \text{ cm}^6 \text{ sec}^{-1} \tag{12.b}$$

Finally, Q_c and Q_e represent contributions due to thermal (molecular) and eddy conduction, respectively:

$$Q_c = \frac{\partial}{\partial z}\left\{\lambda \frac{\partial T}{\partial z}\right\} \tag{13.a}$$

$$Q_e = \frac{\partial}{\partial z}\left\{\rho \, K C p \left(\frac{\partial T}{\partial z} + \Gamma\right)\right\} \tag{13.b}$$

where λ is the thermal (molecular) conductivity, ρ the total density, Cp the specific heat at constant pressure, and Γ is the adiabatic lapse rate $\left(\Gamma = \frac{g}{Cp}\right)$ of the neutral gas.

Numerical Procedure and Boundary Conditions

The equations of continuity and energy balance discussed in the preceding sections are formally all of the same type. They are partial differential equations of first order in t (time) and second order in z (height). The numerical solutions of these equations are obtained by using an implicit finite difference technique, with integration steps $\Delta t = 30$ minutes and $\Delta z = 1$ km. The scheme for solving these equations is basically the same as discussed by Herman and Chandra [1969] and Stubbe [1970].

Equation (5) is relevant only in the lower thermosphere where O and O_2 are minor constituents. In the upper thermosphere O becomes a major constituent as a result of diffusive separation and the distribution of individual constituents and their velocities can be calculated from appropriate equations analogous to (1) and (2), respectively. The upper boundary for equation (5), therefore, has to be chosen at a height where N_2 is still a major constituent and above which level molecular diffusion dominates over eddy diffusion, production and loss terms. We may define this height as a transition height z_t, to distinguish it from a turbopause height.

The boundary conditions at z_t may then be determined from the requirement of the continuity of flux under the specified simplification ($K \ll D_{i3}$):

$$\left[\frac{1}{n_i}\frac{\partial n_i}{\partial z} + \frac{1}{T}\frac{\partial T}{\partial z} + \frac{1}{H_i}\right]_{z=z_t} = \left[\frac{1}{D_{i3}}\left(V_{N_2} - H_i\frac{\partial p_i}{\partial t}\right)\right]_{z=z_t}, \quad (i = 1, 2)$$

If z_t is chosen above 120 km, where the typical values of $D_{i3} \gtrsim 10^8\ cm^2\ sec^{-1}$, the expression on the R.H.S. becomes negligibly small. In this case, the boundary values at z_t may be simply determined from the following condition representative of diffusive equilibrium distribution:

$$\left.\frac{\partial n_i}{\partial z}\right|_{z=z_t} = - \left[n_i \left(\frac{1}{T} \frac{\partial T}{\partial z} + \frac{1}{H_i} \right) \right]_{z=z_t} \tag{14.a}$$

For the lower boundary z_o, we may assume the photochemical terms to dominate over the transport term. The time-dependent boundary conditions for O and O_2 densities in this case may therefore be determined from the following equation:

$$\left.\frac{\partial n_i}{\partial t}\right|_{z=z_o} = (q_i - L_i)_{z=z_o}$$

i.e.,

$$n_i (z_o, t) = n_i (z_o, t - \Delta t) + \Delta t\, (q_i - L_i)_{z_o, t} \tag{14.b}$$

The concept of the transition height z_t is useful only for solving the continuity equations of O and O_2. For solving energy balance equation (eqn. 10), it is not necessary to treat the lower and upper regions of the thermosphere differentially through introduction of a transition height. The upper and lower boundary conditions used in solving equation (10) are as follows:

$$\left.\begin{array}{lll} \text{At} & z = z_u \text{ (450 km):} & \left.\dfrac{\partial T}{\partial z}\right|_{z_u} = 0 \quad , \\ \text{and at} & z = z_o \text{ (85 km):} & T(z_o) = 180°K \end{array}\right\} \tag{14.c}$$

NUMERICAL RESULTS AND DISCUSSION

In presenting the numerical solutions of the energy equation, the choice of solar flux is clearly of critical importance. In a recent paper, Hinteregger [1970] has given a table of spectral distribution of solar EUV flux which he believes to be representative of medium solar activity. The values of EUV flux given in this paper differ considerably from those given by Hinteregger et al. [1965] for reportedly low solar activity and are, on the average, a factor of two less than their earlier values. Hinteregger [1970] also believes that the solar flux in the Schumann-Runge Continuum as tabulated by Hinteregger et al. [1965] is much too high and should be reduced by a factor of 3. Using the solar flux data given by Hinteregger [1970] and the currently accepted range of values of eddy diffusion coefficient we find the exospheric temperature is much too low for the conditions of medium solar activity ($T_\infty < 600°K$). This is in agreement with the conclusion arrived at by Roble and Dickinson [1973]. On the other hand, the use of the solar flux values given by Hinteregger et al. [1965] gives an exospheric temperature much too high for low solar activity ($T_\infty > 1500°K$) consistent with our earlier result [Chandra and Sinha, 1973]. Because of the parametric nature of our present study, we have adopted solar flux values which can accommodate large variations in eddy diffusion coefficient and yet give the temperature variations in a reasonable range. The flux values in EUV range used in this paper are the same as given by Hinteregger [1970]. The flux values in the Schumann-Runge Continuum are taken from Hinteregger et al. [1965] and

the flux values in the Schumann-Runge Band are the same as used by Shimazaki [1971]. The absorption and ionization cross sections are appropriately taken from Hinteregger et al. [1965] and Shimazaki [1971].

The eddy coefficient (K-model) as defined by eqn. (9) has several parameters which affect the altitude distribution of the eddy diffusion in the lower thermosphere. For the purpose of parametric studies, however, here we consider only the variations in K_M and z_m, which are the main parameters describing the general features of the eddy turbulence in this region. The overall changes in K(z) associated with the changes in K_M and z_m are shown in Figures (1a) and (1b), respectively. In both these plots, the values of S_1, S_2, S_3 and K_o as defined in eqn. (9) are fixed as follows:

$$S_1 = 0.05\,\text{km}^{-2},\ S_2 = 0.03\,\text{km}^{-2},\ S_3 = 0.07\,\text{km}^{-1},\ K_o = 10^6\,\text{cm}^2\,\text{sec}^{-1}.$$

For the case depicted in Figure (1a), z_m has a fixed value of 110 km and K_M varies from 10^6 to 10^8 cm^2 sec^{-1}. The effect of this change on K(z) is most noticeable in the immediate vicinity of the height of the maximum z_m. The various profiles merge into a single profile about 15 km below z_m. In the altitude region above z_m, K diminishes to a low value well before the transition height z_t (=130 km) for all the four K_M values shown. Clearly here higher K_M value corresponds to higher K-value at all altitudes above 95 km. In contrast to this situation the variation of z_m from 120 to 105 km with K_M fixed at 5×10^7 cm^2 sec^{-1} (Fig. 1b), affects the K-values differently in different altitude regions. The lowering of z_m yields a higher K-value in the lower altitude region and lower K-values in the higher altitude region.

As a direct consequence of the altitude variation in K(z), the eddy term associated with the adiabatic lapse rate in equation (13.b) acts as a heat source or sink in the various altitude regions. Denoting this term by Q_{ad} we may write from eqn. (13.b).

$$Q_{ad} = g \rho K \left[\frac{1}{K} \frac{\partial K}{\partial z} + \frac{1}{\rho} \frac{\partial \rho}{\partial z} \right] \tag{15}$$

Since

$$\frac{1}{\rho} \frac{\partial \rho}{\partial z} \approx -\frac{1}{H}, \text{ and } \frac{1}{K} \frac{\partial K}{\partial z} < 0 \text{ for } z > z_m,$$

Q_{ad} is always negative in the altitude region above the height z_m and hence is an effective heat sink in this region. In the altitude region below z_m, $\frac{\partial K}{\partial z} > 0$ and Q_{ad} may act either as a heat source or sink depending upon the sign of $\left[\frac{1}{K} \frac{\partial K}{\partial z} + \frac{1}{\rho} \frac{\partial \rho}{\partial z} \right]$. The variations in K(z), therefore, results in the increase or decrease of the neutral temperature in the various regions of the thermosphere depending on the nature of Q_{ad}. The eddy related changes in the neutral composition are even more difficult to describe in general, since the changes in K(z) affect the effective scale height of the minor constituents both directly (via mixing) and indirectly (via neutral temperature). Some of the intricacies of these interacting processes can be described in detail only through time-dependent numerical solutions presented in the following sections.

Variation of K_M (z_m Constant)

The temperature and atomic oxygen density profiles corresponding to the K-models discussed in Figure (1a) are shown in Figure (2) for the local noon

condition. The numerical results are shown here only up to about 200 km since there are no basic changes in the trends of the various profiles above this altitude region. The temperature in all the cases increases continuously with altitude until it becomes nearly isothermal (at about 300 km), and the atomic oxygen density, in accordance with the diffusive equilibrium distribution, falls off exponentially at a rate determined by its scale height. Both the temperature and atomic oxygen profiles have several interesting features in the altitude region of 85-200 km, which are summarized as follows:

(1) The temperature varies inversely as K(z) according to the qualitative deductions discussed in the preceding section.

(2) The temperature, in general, does not increase monotonically with altitude, but instead goes through a maximum and a minimum in the altitude region of 95-120 km. The change from maximum to minimum depends upon the values of K(z) and is attributed to the heating and cooling associated with the adiabatic lapse rate given by eqn. (15). Low K_M value, however, yields a monotonic temperature profile, as exemplified by $K_M = 10^6 cm^2 sec^{-1}$ curve.

(3) The variations in K_M do not have any significant effect on [O] in the altitude region below 100 km, since this region is basically under photochemical control. In the altitude region above 100 km, [O] decreases with increase in K(z). The effect of increasing K(z) is to decrease the effective scale height of atomic oxygen directly (through increased mixing) and indirectly (through a reduced neutral temperature).

The diurnal variation of the exospheric temperatures (taken at 450 km) for representative values of K_M in this series is represented in Fig. (3). As seen from this plot, the change in K_M affects both the diurnal amplitude and the absolute values of temperature but not the basic diurnal characteristic with respect to the times of the maximum and minimum. A decrease in K_M from 10^8 to 10^6 cm^2-sec^{-1} results in increase of the exospheric temperature by about 50%. A change of this magnitude is comparable to the changes associated with the solar activity over a solar cycle.

In presenting the numerical results of Figs. (2) and (3) the eddy diffusion models were assumed to be independent of time, since their diurnal variations are not known with any certainty. In the spirit of parametric studies, we have investigated the effect of varying K_M over a diurnal time scale. The time varying models of K_M have little effect on the altitude and diurnal characteristics of temperature and densities, though they do change their absolute values. A decrease in K_M by an order of magnitude from day to night, for example, increases the exospheric temperature by about 20% without changing its basic altitude characteristics.

Variations of z_m (K_M Constant)

The altitude variations in temperatures and [O] as a result of changing z_m are quite different from those due to the variations in K_M. Fig. (4) is a typical example of results for changing z_m from 120 to 105 km with $K_M = 5 \times 10^7 cm^2 sec^{-1}$ (Fig. 1b) during the day with one order of magnitude reduction in K-value during the night. (As mentioned already, the results for $K_M = 5 \times 10^7 cm^2 sec^{-1}$ throughout the day and the night are basically similar except for some differences in the absolute magnitudes.) Unlike the changes associated with K_M, the changes in z_m have different effects in the upper and the lower thermospheric regions. The reduction in z_m results in the increase of the temperature in the higher altitude region and decrease in the lower altitude as a direct consequence of the redistribution of K(z) and hence of the adiabatic lapse rate related cooling (heating) rate (eqn. 15).

The increase in K(z) in the lower thermosphere reduces the atomic oxygen in this region. However, this reduction does not persist at higher altitudes because of the compensating effects from the increased temperature. The crossover in temperature from a lower value in the lower thermosphere to a higher value in the upper thermosphere produces a similar effect on atomic oxygen and other neutral constituents through corresponding changes in their scale heights. More specifically, by lowering of z_m from 120 to 105 km, the atomic oxygen density decreases in the lower thermosphere due to reduction in its scale height. This decrease persists up to about 200-350 km, above which the pattern

gradually changes to increase in [O]. This kind of cross-over occurs in other constituents also. The altitude at which the crossover occurs for a specific change in z_m depends on the constituent. For the heavier constituents, (O_2, N_2) and also for the total density (Fig. 5), the crossover takes place in a much lower altitude range. An example of the changes in O/N_2 and O/O_2 associated with the change in z_m is also shown in Fig. (5). The effect of decreasing z_m, as seen in this plot, is to decrease O/O_2 and O/N_2. This decrease is much more apparent in the higher altitude region ($z > 160$ km) and is a direct consequence of the differences in the scale heights of the three constituents above the transition height.

SUMMARY AND CONCLUDING REMARKS

In this paper, we have attempted to emphasize the importance of coupling between the lower and the upper thermosphere and the specific effects of eddy turbulence on the thermal structure and the neutral composition in the thermospheric region. Because of the complex nature of interactions of the various processes, we numerically solved the time-dependent equations of continuity and energy balance of the relevant neutral constituents in a self-consistent manner and without making ad-hoc assumptions about the composition and temperature changes at the turbopause.

Our study relating to the effects of eddy turbulence covered a wide range of variations in the eddy diffusion coefficient without relating these variations to any specific geophysical causes. The numerical results presented in this paper correspond to the geographic equator, though the qualitative inferences drawn from

these calculations are quite general and are applicable to the high and mid-latitude regions. Notwithstanding the parametric nature of this study, it seems reasonable to draw certain conclusions regarding the geophysical implications of the types of results obtained. Variation in the magnitude K_M and altitude z_m of eddy coefficient may have important implications in regard to seasonal variation of the neutral temperature and composition [Kockarts, 1972].

In particular, the temperature and composition changes associated with variations in z_m are analogous to those actually observed during a geomagnetic storm at middle and high latitudes. A large reduction in O/N_2 ratio, together with increase in exospheric temperature, are among the characteristic changes in the thermosphere during a magnetic storm. [Jacchia et al., 1966; Blamont and Luton, 1972; Reber et al., 1973]. We observe that lowering of z_m induces precisely these types of effects. Thus we are led to the tentative conclusion that during the main phase of a geomagnetic storm the turbulence in the lower thermosphere is shifted to lower altitudes. This is physically plausible in view of low altitude deposition of energy in the auroral region due to energetic particle precipitation and Joule dissipation during magnetic storms. Subsequent transport of energy to the mid-latitude regions is expected to yield enhanced turbulence at lower altitudes, which is schematically analogous to lowering of the z_m-value.

Obviously, by the very nature of the process, eddy turbulence is generally a highly random and variable phenomenon. Variations in the eddy diffusion

coefficient, a purely phenomenological parameter, are to be expected with changes in the geophysical conditions, such as tropospheric and stratospheric disturbances of meteorological nature, gravity and tidal waves, local solar heating, magnetospheric and field-aligned perturbations, etc. The effect of such variations on thermospheric temperature and composition can, in principle, be incorporated in model calculations via appropriate modifications in the eddy coefficient, as illustrated here. It is, therefore, evident that inclusion of the eddy turbulence as a basic physical process is important for constructing a coherent model of the upper atmosphere and the ionosphere.

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FIGURE CAPTIONS

Figure 1-a. The altitude profiles of K(z) for K_M varying from 10^6 to 10^8 cm^2-sec^{-1}.

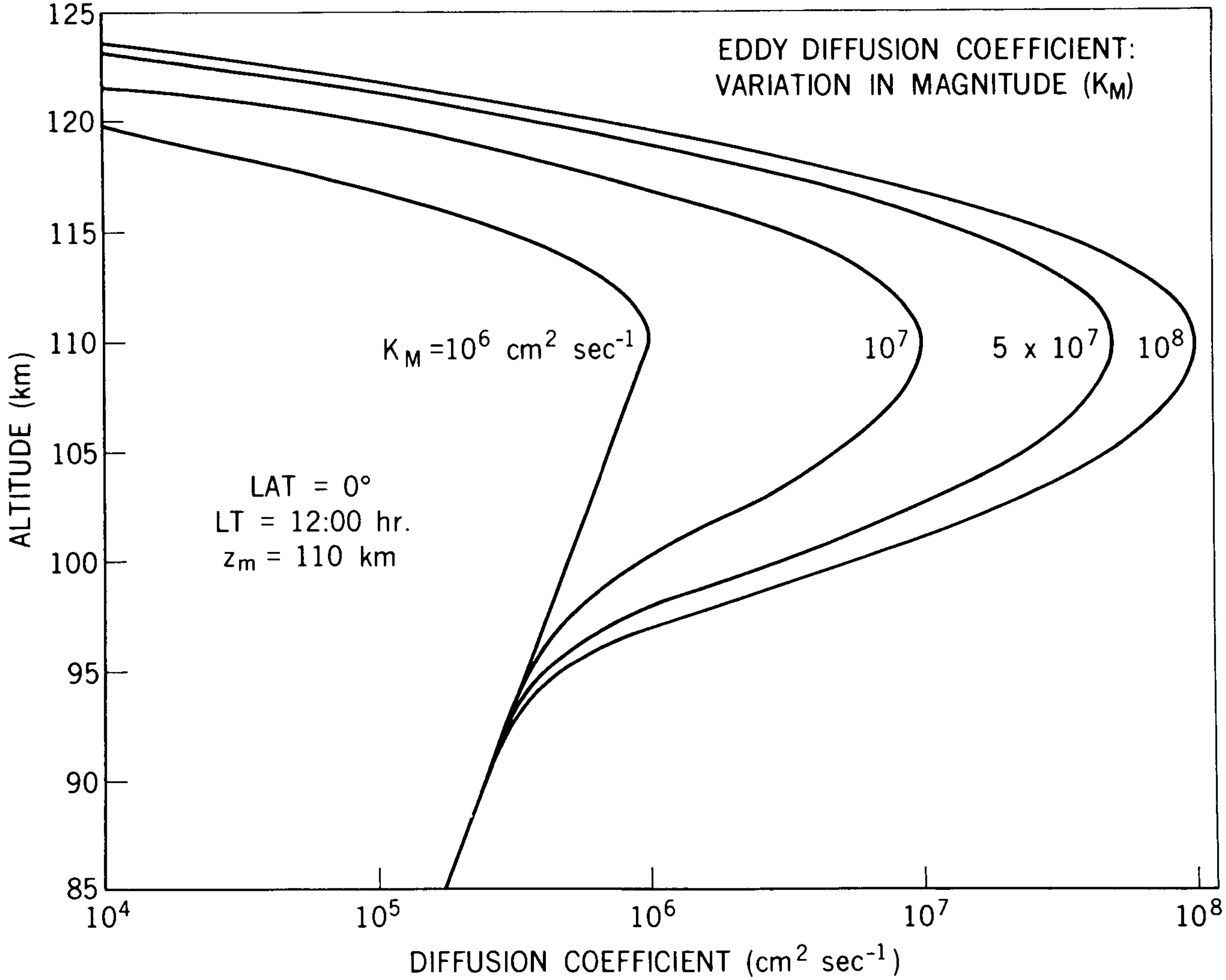
Figure 1-b. The altitude profiles of K(z) for z_m varying from 105 to 120 km.

Figure 2. The altitude profiles of temperature and atomic oxygen density for K-models shown in Figure 1-a.

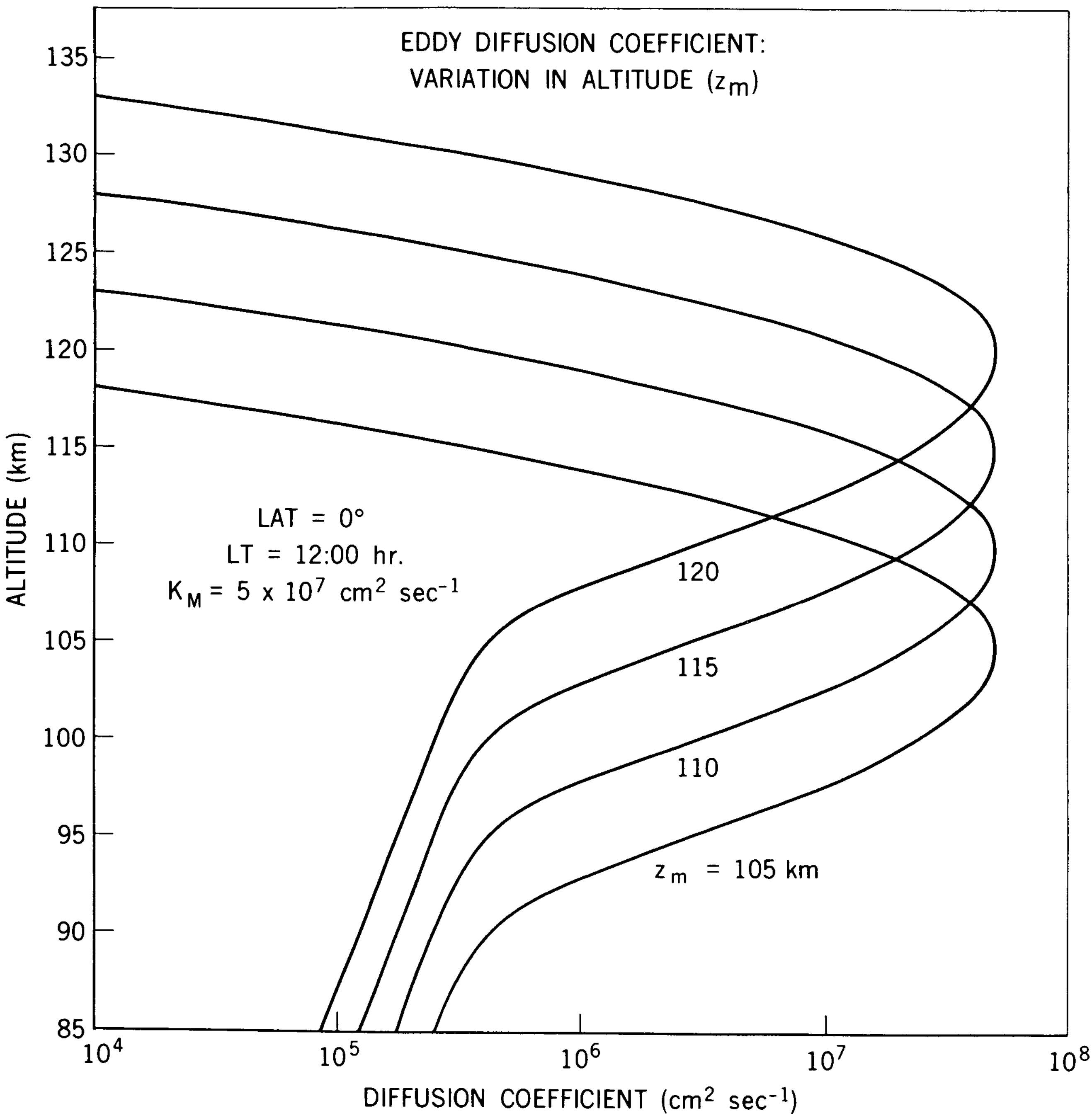
Figure 3. The diurnal variations in exospheric temperature for varying values of K_M.

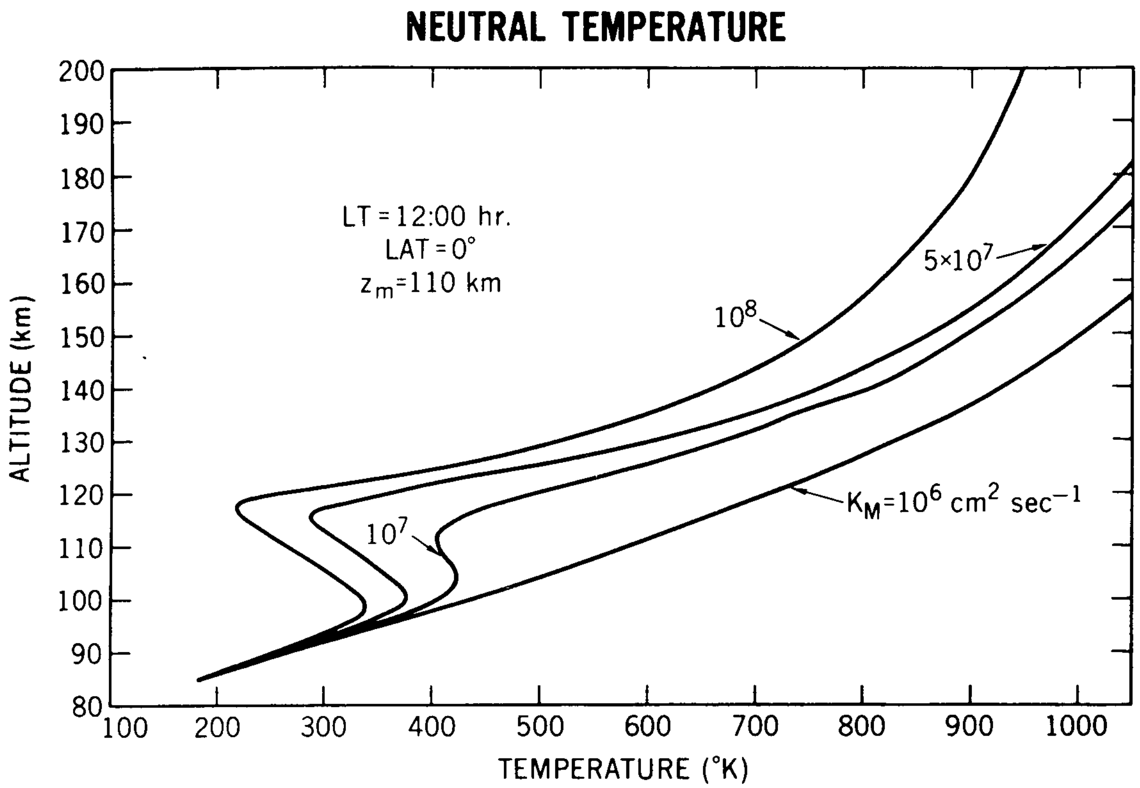
Figure 4. The altitude profiles of temperature and atomic oxygen density for K-models shown in Figure 1-b.

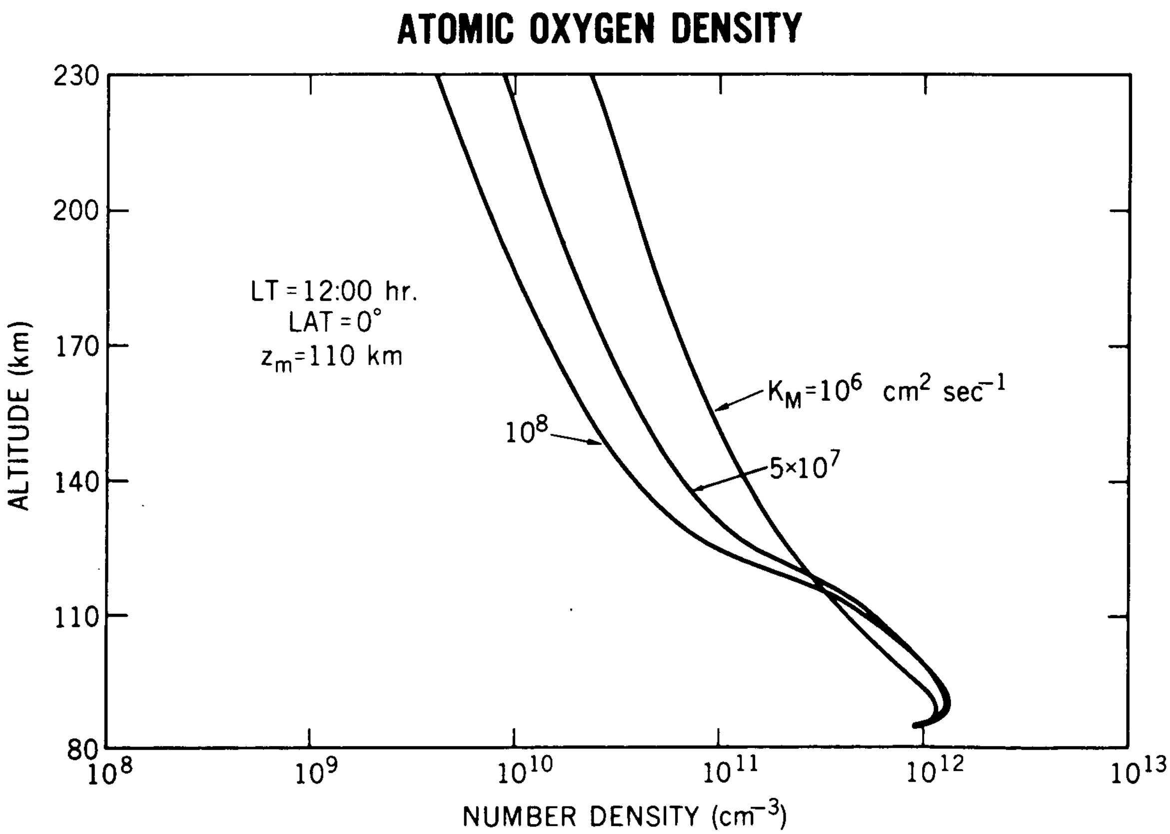
Figure 5. The altitude profiles of O/N_2, O/O_2 and the total density (mass) for K-models of Figure 1-b.


EDDY DIFFUSION COEFFICIENT:
VARIATION IN MAGNITUDE (K_M)
$K_M = 10^6$ cm^2 sec^{-1}
10^7
5×10^7
10^8
LAT = 0°
LT = 12:00 hr.
z_m = 110 km
ALTITUDE (km)
125
120
115
110
105
100
95
90
85
10^4
10^5
10^6
10^7
10^8
DIFFUSION COEFFICIENT (cm^2 sec^{-1})

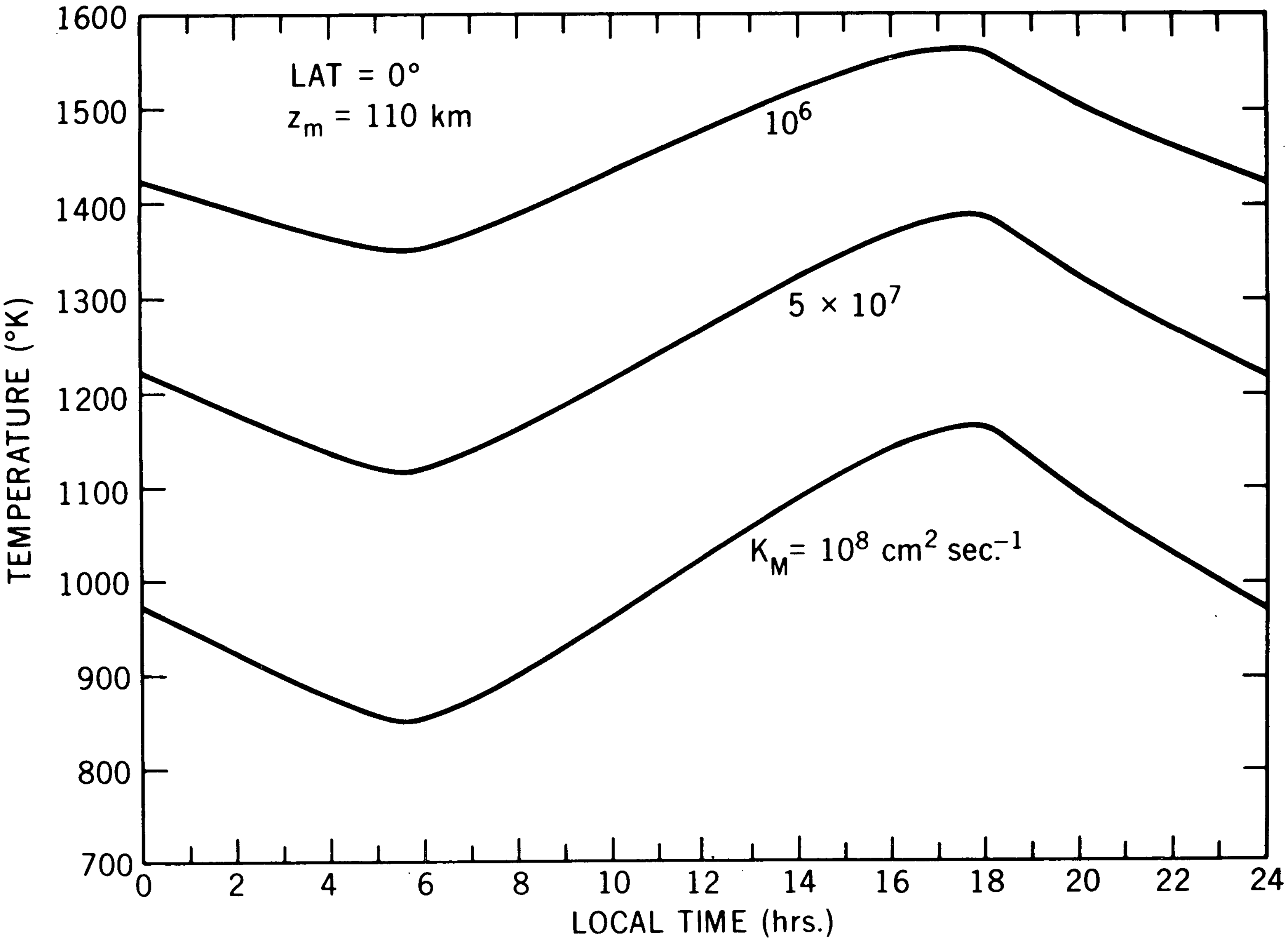
27

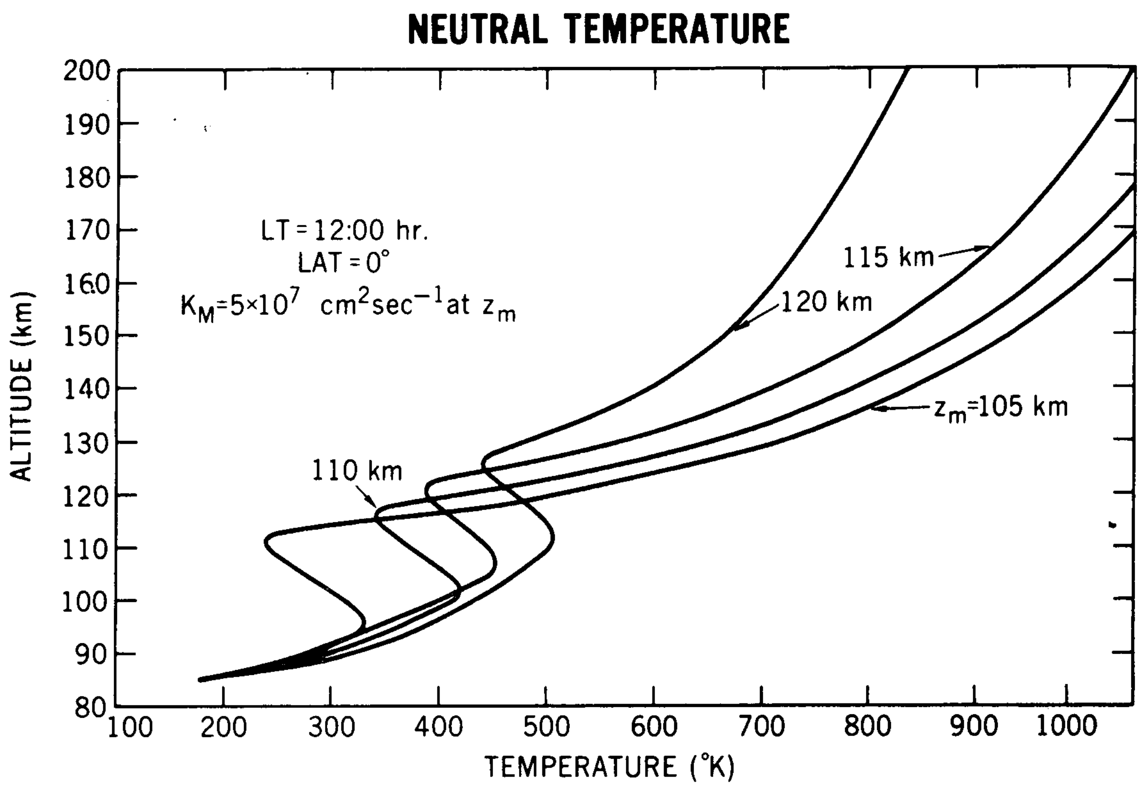

EDDY DIFFUSION COEFFICIENT:
VARIATION IN ALTITUDE (z_m)
LAT = 0°
LT = 12:00 hr.
$K_M = 5 \times 10^7$ cm^2 sec^{-1}
120
115
110
z_m = 105 km
ALTITUDE (km)
135
130
125
120
115
110
105
100
95
90
85
10^4
10^5
10^6
10^7
10^8
DIFFUSION COEFFICIENT (cm^2 sec^{-1})

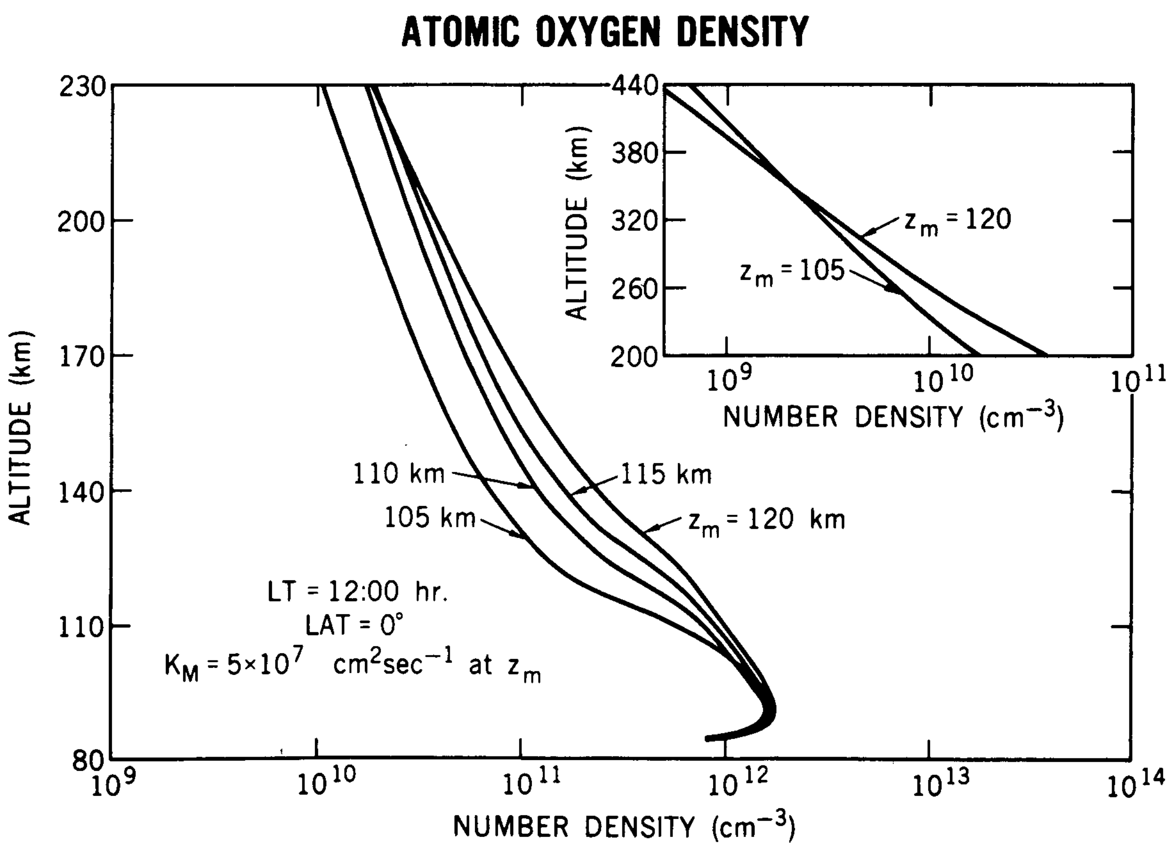

NEUTRAL TEMPERATURE
LT = 12:00 hr.
LAT = 0°
z_m = 110 km
10^8
5×10^7
10^7
K_M = 10^6 cm^2 sec^-1
ALTITUDE (km)
TEMPERATURE (°K)
200
190
180
170
160
150
140
130
120
110
100
90
80
100
200
300
400
500
600
700
800
900
1000


ATOMIC OXYGEN DENSITY
LT = 12:00 hr.
LAT = 0°
z_m = 110 km
K_M = 10^6 cm^2 sec^-1
10^8
5×10^7
ALTITUDE (km)
NUMBER DENSITY (cm^-3)
230
200
170
140
110
80
10^8
10^9
10^10
10^11
10^12
10^13

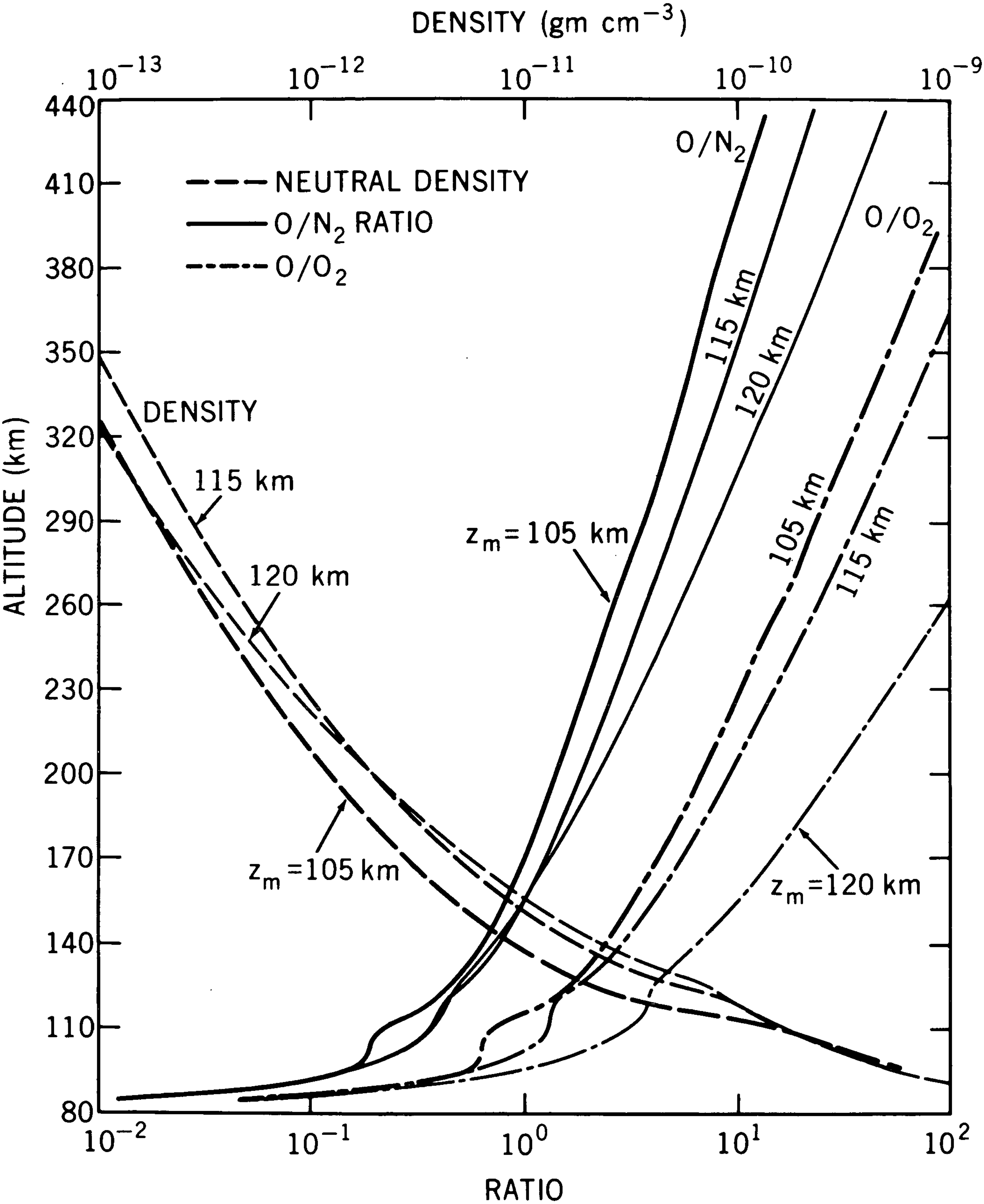
EXOSPHERIC TEMPERATURE: DIURNAL VARIATION




NEUTRAL TEMPERATURE
LT = 12:00 hr.
LAT = 0°
$K_M = 5\times10^7$ cm^{2}sec^{-1} at z_m
115 km
120 km
110 km
z_m=105 km
ALTITUDE (km)
TEMPERATURE (°K)


ATOMIC OXYGEN DENSITY
z_m = 120
z_m = 105
ALTITUDE (km)
NUMBER DENSITY (cm^{-3})
110 km
115 km
105 km
z_m = 120 km
LT = 12:00 hr.
LAT = 0°
$K_M = 5\times10^7$ cm^{2}sec^{-1} at z_m
ALTITUDE (km)
NUMBER DENSITY (cm^{-3})

32


DENSITY (gm cm^{-3})
10^{-13}
10^{-12}
10^{-11}
10^{-10}
10^{-9}
440
410
380
350
320
290
260
230
200
170
140
110
80
ALTITUDE (km)
NEUTRAL DENSITY
O/N$_2$ RATIO
O/O$_2$
O/N$_2$
O/O$_2$
115 km
120 km
DENSITY
115 km
120 km
z_m=105 km
105 km
115 km
z_m=105 km
z_m=120 km
10^{-2}
10^{-1}
10^{0}
10^{1}
10^{2}
RATIO